HYDROLOGICAL PROCESSES
Hydrol. Process. **22**, 1885–1894 (2008)
Published online 4 September 2007 in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/hyp.6773

Hyporheic and total transient storage in small, sand-bed streams

John M. Stofleth,[1] F. Douglas Shields Jr[2*,†] and Garey A. Fox[3]

[1] *Philip Williams & Associates, Sacramento, CA, USA*
[2] *USDA-ARS National Sedimentation Laboratory, Oxford, MS, USA*
[3] *Oklahoma State University, Department of Biosystems and Agricultural Engineering, Stillwater, OK, USA*

Abstract:

Key processes in stream ecosystems are linked to hydraulic retention, which is the departure of stream flow from ideal 'plug flow', and reflects fluid movement through surface and hyporheic storage zones. Most existing information about hyporheic exchange is based on flume studies or field measurements in relatively steep streams with beds coarser than sand. Stream tracer studies may be used to quantify overall hydraulic retention, but disaggregation of surface and hyporheic retention remains difficult. A stream tracer approach was used to compute the rates at which stream water is exchanged with water in storage zones (total storage) in short reaches of two small, sand-bed streams under free and obstructed flow conditions. Tracer curves were fit to the one-dimensional transport with inflow storage model OTIS-P. Networks of piezometers were used to measure specific discharge between the stream and the groundwater. In the sand-bed streams studied, parameters describing total retention were in the upper 50% of data compiled from the literature, most of which represented streams with beds coarser than sand. However, hyporheic storage was an insignificant component of total hydraulic retention, representing only 0·01–0·49% of total exchange, and this fraction did not increase after installation of flow obstructions. Total retention did not vary systematically with bed material size, but increased 50–100% following flow obstruction. Removal of roughness elements, such as large wood and debris dams, is detrimental to processes dependent upon transient storage in small, sand-bed streams.

KEY WORDS hydraulic retention; streams; sand; hyporheic zones; tracers; large wood; piezometers; transport

Received 18 September 2006; Accepted 20 March 2007

INTRODUCTION

The hyporheic zone is a porous layer of the streambed affected by small-scale exchange between stream water and shallow groundwater (Harvey and Wagner, 2000). Typically, water flowing in the stream channel flows into the subsurface materials of the streambed (hyporheic zone) and then returns to the stream (Bencala, 2005). Hyporheic zones support intense biological (e.g. Hendricks, 1993) and biogeochemical activity such as nutrient transformations (Triska *et al.*, 1993). Hyporheic zones are a subset of a group of features termed 'transient storage zones' in which water velocity is slower than that of the main advective flow (Bencala and Walters, 1983). Transient storage zones may be below the bed (hyporheic storage) or in the stream. In-stream storage zones include topographic features (e.g. side pools), and regions influenced by large wood or riparian vegetation (Chapra and Wilcock, 2000; Harvey *et al.*, 2003; Ensign and Doyle, 2005). The presence of obstructions such as beaver dams or debris formations that create surface storage may also enhance exchange with subsurface storage zones (Mutz and Rohde, 2003; Lautz *et al.*, 2006). The attendant increase in mean residence time due to these above- and below-streambed surface storage zones is referred to as hydraulic retention or total storage. Biological and geochemical processes that occur in storage zones are responsible for trapping and processing nutrients and other vital ecological processes (Munn and Meyer, 1990; Boulton, 2000; Hall *et al.*, 2002). Previous research has shown that hydraulic retention, flow resistance, nutrient uptake, and retention of particulate organic carbon are inversely related to flow, and directly related to the complexity of channel morphology (Kasahara and Hill, 2006; Kasahara and Wondzell, 2003), and the loadings of large wood (D'Angelo *et al.*, 1993; Shields and Gippel, 1995; Mutz and Rohde, 2003; Stofleth *et al.*, 2004; Ensign and Doyle, 2005) and leaf litter (Jin and Ward, 2005).

Hydraulic retention (or total storage) has been quantified by applying data from conservative tracer studies to inverse solute transport models (Bencala and Walters, 1983), principally OTIS-P, a one-dimensional transport and storage model with groundwater inflow (Runkel, 1998). OTIS-P models the stream as a two-compartment system composed of the main channel and the storage zone. The conceptual storage zone accounts for all types of non-ideal flow within the reach, whether it is due to surficial 'dead zones', hyporheic storage, or flow non-uniformity (Figure 1). Transport in the main channel is governed by advection and dispersion, with solute transport between the main channel and the storage zone

* Correspondence to: F. Douglas Shields, USDA-ARS National Sedimentation Laboratory, PO Box 1157, Oxford, MS 38655-1157, USA. E-mail: dshields@msa-oxford.ars.usda.gov
† The contribution of F. Douglas Shields was prepared as part of his official duties as a United States Federal Government employee.

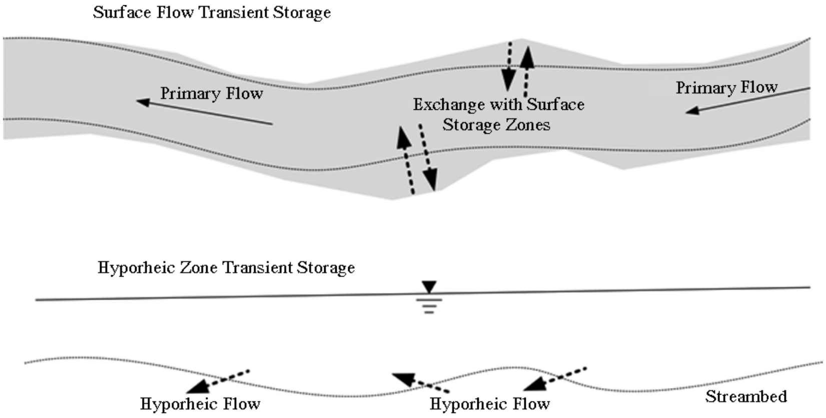

Figure 1. Schematic of transient storage concepts. Upper sketch shows plan view of stream with exchange with surface storage zones. Lower sketch is stream profile, showing exchange with hyporheic zone. Total transient storage = surface flow transient storage + hyporheic zone transient storage. Adapted from Runkel and Bencala (1995)

proportional to a concentration gradient. There is no net movement of solute in the storage compartment, which is treated as a completely mixed dead zone (Bencala and Walters, 1983). OTIS-P fits the following equations to the observed tracer curves (Runkel and Broshears, 1991):

$$\begin{aligned}\frac{\partial C}{\partial t} &= -\frac{Q}{A}\frac{\partial C}{\partial x} + \frac{1}{A}\frac{\partial}{\partial x}\left(AD\frac{\partial C}{\partial x}\right) \\ &\quad + \frac{q_{\mathrm{LIN}}}{A}(C_{\mathrm{L}} - C) + \alpha(C_{\mathrm{s}} - C)\frac{\mathrm{d}C_{\mathrm{s}}}{\mathrm{d}t} \\ &= \alpha\frac{A}{A_{\mathrm{s}}}(C - C_{\mathrm{s}}) \qquad (1)\end{aligned}$$

where C [ML^{-3}] is the stream solute concentration, A [L^2] is the cross-sectional area of the stream channel, Q [$\mathrm{L}^3\mathrm{T}$] is the steam discharge, D [$\mathrm{L}^2\mathrm{T}^{-1}$] is the stream dispersion coefficient, q_{LIN} [T^{-1}] is lateral inflow rate, C_{L} [ML^{-3}] is the solute concentration in lateral inflow, α [T^{-1}] is the storage zone exchange coefficient, C_{s} [ML^{-3}] is the storage zone solute concentration, and A_{s} [L^2] is the storage zone cross-sectional area.

Differentiation between surface and hyporheic zone storage has been difficult (Harvey and Wagner, 2000). Differentiation is of interest for several reasons (Jin and Ward, 2005), one of which is the suggestion by some that the potential for nutrient processing in the hyporheic zone is greater than that of surface storage zones (Triska *et al.*, 1989), and the fact that nutrient uptake is only weakly related to total transient storage (Mulholland *et al.*, 1997; Hall *et al.*, 2002). Unfortunately, research in hyporheic functions and processes is complicated by the relative inaccessibility and complexity of the hyporheic zone (Palmer, 1993). Many of the existing studies do not report vertical hydraulic gradient (VHG) at the streambed, hydraulic conductivity, or even bed material grain size (e.g. Broshears *et al.*, 1996; Hart *et al.*, 1999), information that would aid in differentiating surface and hyporheic storage. Harvey and Wagner (2000) surveyed existing stream tracer studies and concluded that the problem of distinguishing between hyporheic and surface storage needed more research, particularly for larger, faster flowing streams. They called for new experiments 'designed to isolate and distinguish hyporheic and in-stream storage processes'. Gooseff *et al.* (2005) compared tracer breakthrough curves in two streams: a cascade–pool-type bedrock reach with no hyporheic zone and an alluvial reach with significant hyporheic exchange. They suggested that in-stream transient storage matched exponential residence time distributions, whereas hyporheic storage matched power law residence time distributions. However, they did not attempt to separate the magnitude of surface and hyporheic storage.

The burgeoning literature on the biology and geochemistry of hyporheic zones is heavily weighted toward mountain streams with beds of gravel and coarser materials. Most existing information on hydraulic retention has been obtained from tracer studies in montane, coarse-bed streams, and much less information is available for streams with sandy or silty beds, particularly smaller streams of the eastern USA with disturbed channels in agricultural or mixed-cover watersheds (Harvey and Wagner, 2000; Battin *et al.*, 2003; Cleven and Meyer, 2003; Hauer *et al.*, 2003). It is generally assumed that hyporheic storage is a smaller component of total storage in sand-bed streams than in gravel/cobble/boulder streams, but data are lacking (D'Angelo *et al.*, 1993; Harvey and Wagner, 2000; Jin and Ward, 2005). An exception is work by Jin and Ward (2005), who measured the flow cross-sectional area at nine transects within their study reach of a small sand-bed stream and compared the average of these measurements, A_{c}, with average values of A output by OTIS-P. They reasoned that the difference between A_{c} and the computed area A represented surface storage. On average, A_{c} was about 1·6 times as large as A, but the two quantities were nearly equal when A_{c} was limited to riffle cross-sections, where little surface storage was found. Thus, they concluded that their stream had little hyporheic storage relative to surface storage. This is consistent with the findings of others (Hancock *et al.*, 2001, Boulton *et al.*, 2002).

In this study we quantify hyporheic and total storage exchange in sand-bed streams differing in bed material size and channel morphology. In addition, we

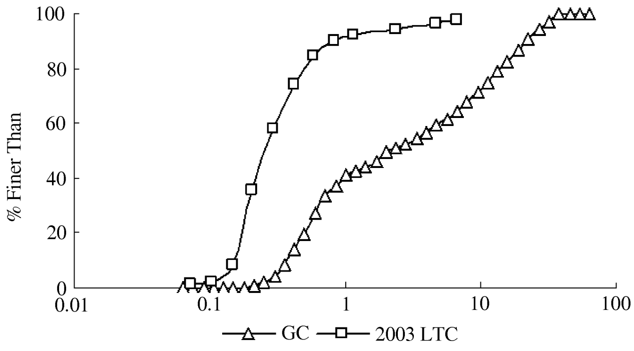

Figure 2. Bed material size gradations for GC and LTC

examine changes in transient storage parameters associated with addition of large wood obstructions. Our hypotheses were that hydraulic retention would be positively related to the presence of flow obstructions and bed material size. Limited research has been performed on stream–hyporheic zone interactions in sand-bed, lowland streams. The findings described below thus potentially impact environmental management issues associated with biogeochemistry of lowland stream beds.

SITE DESCRIPTION

The two sand-bed streams investigated are located in north-central Mississippi, USA. The streams are similar in catchment size and overall form, but differ in bed material size and bed morphology. Little Topashaw Creek (LTC, http://www.ars.usda.gov/Research/docs.htm?docid=5526, accessed 1 August 2006) drains a fourth-order, 37 km^2 north-central Mississippi watershed and has cultivated valley floors and forested hillslopes. Floodplain stratigraphy is characterized by dispersive silt and clay soils over sand that overlies consolidated cohesive material. Channels upstream and downstream from the study reach were straightened *c*. 1913 and downstream reaches were again channelized in 1967. These activities have induced massive incision and widening throughout the reach. Bed materials are uniform, comprised primarily of medium sand ($D_{50} = 0{\cdot}27$ mm), and channel slope is mild (Figure 2, Table I). Streams are characterized by flashy hydrology and elevated sediment loads. In our study reach, an impermeable restrictive layer of micaceous shale occurred about 1 m below the streambed (Adams, 2000). The channel is tortuous, with an average sinuosity of 2·1 and with average channel width and depth of 33·3 m and 3·6 m respectively.

Goodwin Creek (GC, http://www.ars.usda.gov/Business/docs.htm?docid=5040, accessed 1 August 2006) drains a fourth-order, 21 km^2 northwest Mississippi watershed located along the bluffline of the Mississippi River Valley. The main channel of GC was channelized (straightened) prior to about 1940 and experienced severe incision between 1960 and 1980. Channel width ranged from 20 to 70 m and depth from 4 to 5 m. Bed material was a bimodal mixture of sand and gravel ($D_{50} = 2{\cdot}1$ mm), and channel slopes were mild (Figure 2, Table I). Although the study reach was flanked on one side by cotton fields, watershed land use was primarily forest, pasture, and fallow lands.

METHODS

Tracer tests

Injections of a conservative tracer were used to characterize the total (surface and hyporheic) retention at LTC and GC (Table II). To minimize uncertainty in parameters derived using OTIS-P, the length of the study reach was selected so that the Damkohler number DaI $\cong$ 1 (Harvey and Wagner, 2000). Preliminary estimates of velocity u (m s^{-1}), cross-sectional area A (m^2), storage zone area A_s (m^2), and α (s^{-1}) exchange coefficient were used in the pre-experiment computations of DaI.

For each test, a solution of sodium chloride (NaCl) with a concentration of 192 or 240 g l^{-1} was continuously injected at a constant rate of 540 ml min^{-1} uniformly across the flow area 30 to 50 m (to ensure adequate mixing) upstream of the study reach during baseflow conditions. Specific electrical conductance was measured using a YSI model 85 meter at the upstream and downstream ends of the reach for 2 to 4 h at intervals designed to characterize the breakthrough curve (Table II). The injection rate and concentration of NaCl were selected based on stream discharge to increase specific electrical conductance above background levels. After initial tracer tests were run, the effects of natural and artificial flow obstructions on hydraulic retention were investigated at GC and LTC respectively (Table II). At GC a beaver constructed a small dam (~0·35 m high spanning about 80% of flow width) within the study reach. Tracer tests were run with the beaver dam in place on 30 July 2004 and 13 August 2004. At LTC, an artificial obstruction was installed to mimic the effects of a large wood formation (underflow jam; Wallerstein *et al*., 1997). The obstruction consisted of a 76·2 cm diameter corrugated PVC pipe, positioned lengthwise across the

Table I. General hydrologic characteristics of study site, including the average water width, depth and velocity during experiments

Stream	Slope (m m^{-1})	Flow width (m)	Flow depth (m)	Bed material size D_{50} (mm)	Velocity (m s^{-1})	Discharge (l s^{-1})	Watershed area (km^2)	CEM stage[a]
GC	0·001	4·79	0·12	2·09	0·08	23–53	21	V
LTC	0·002	4·75	0·06	0·27	0·15	35–51	37	VI

[a] Conceptual model of incised channel evolution proposed by Simon (1989).

Table II. Dates and durations for tracer injection experiments. Duration includes time from first to last observation. Discharge and length are of the study reach during the experiments

Stream	Date (2004)	Duration (min)	Obstruction	Discharge (l s^{-1})	Length (m)
GC	26 May	182	No	39	54
	7 June	260	No	53	53
	14 July	261	No	53	64
	30 July	151	Full beaver dam	23	63
	13 August	258	Partial beaver dam	25	63
LTC	22 July	311	No	51	65
	4 August	210	Artificial underflow jam	36	65

stream and partially filled with concrete. The tracer test was performed 6 days after obstruction installation.

Piezometric measurements

For each tracer test, networks of piezometers were installed at each site to characterize the hyporheic exchange at the streambed surface. The piezometers were constructed of open-ended 15·2 cm diameter PVC pipes and were inserted in the streambed to a depth of 25 cm. No attempt was made to install piezometers at different depths in the streambed. The 25 cm depth was chosen as a balance between installation (i.e. difficult to install the piezometers at shallower depths without washout of sediment) and quantifying surface exchange processes (i.e. installing piezometers to deeper depths may have masked exchange at the streambed surface). Piezometers were placed in the thalweg of the study reach at a spacing of 5 m following a pilot study with closer spacing. Each piezometer was assumed to represent an equal area of the streambed (i.e. 2·5 m upstream and 2·5 m downstream of each piezometer), which was realistic for these lowland, sand-bed streams because flow and substrate conditions changed gradually in the streamwise direction. At each piezometer, VHG was measured (Baxter *et al.*, 2003) and falling-head tests (Landon *et al.*, 2001) were performed following each tracer test. VHG was determined from the difference in interior and exterior water surface elevations at each piezometer. The exterior water surface elevation was measured at the water's edge adjacent to the piezometer using a total station. The interior elevation was determined by surveying the top of the pipe with the total station and then measuring the distance from the top of the pipe to the interior water surface with a tape or metre stick. Falling-head tests were performed by adding water to each piezometer to impose an additional 10 to 20 cm head on the sediments inside the pipe. Water surface displacement was then recorded every minute over a 5 min period, which appeared adequate to characterize test results based on the resulting smooth, linear curves of water level versus time. Values of vertical hydraulic conductivity were computed using Darcy's solution for falling-head tests (Landon *et al.*, 2001; Fox, 2004).

Hydraulic measurements

During each tracer test, discharge was measured at the upstream and downstream ends of each study reach using dilution gauging techniques (Table II). Discharge was also measured using a wading rod and an electromagnetic current meter (area–velocity method) within the study reach for quality control. A thalweg profile and selected points along the water's edge were surveyed using a total station and these data were used to compute mean bed slope and flow width.

Metrics of total storage

The model OTIS-P was used to compute values for α, A_s, A and D for each tracer test and then these values were used in various metrics to quantify total storage, such as the hydraulic retention factor and the fraction of median travel time due to transient storage. Model outputs represented the suite of parameter values that produced a 'best fit' to the observed specific conductance versus time curve (Figure 3) (Bencala and Walters, 1983; D'Angelo *et al.*, 1993). Surface flow hydraulics were characterized by computing the Darcy friction factor f for each reach during each tracer test as follows (Jin and Ward, 2005):

$$f = \frac{8gds}{u^2} \tag{2}$$

where g is acceleration due to gravity, d is average flow depth, s is streambed slope, and u is reach mean flow velocity. Flow depth was computed by dividing the flow cross-sectional area A obtained from OTIS-P by the mean flow width, and velocity was computed by $u = Q/A$.

The hydraulic retention factor R_h (Morrice *et al.*, 1997) and the fraction of the median travel time due to transient storage F_{med} were computed as follows (Runkel, 2002):

$$R_h = \frac{T_{sto}}{L_s} = \frac{A_s}{Q} \text{ where } T_{sto} = \frac{A_s}{\alpha A} \text{ and } L_s = \frac{u}{\alpha} \tag{3}$$

$$F_{med} = (1 - e^{-L\alpha/u})\frac{A_s}{A + A_s} \tag{4}$$

where T_{sto} is the storage zone residence time and L_s is the distance travelled before entering the storage zone.

F_{med} reflects the interaction between advective velocity and transient storage. For the purposes of comparing values of F_{med} from different sites and experiments, Runkel (2002) suggests that a reach length $L = 200$ m be used in Equation (3). All values reported herein are for F_{med200}.

Distributions of hydraulic conductivity for the two study streams were compared using a box-and-whisker

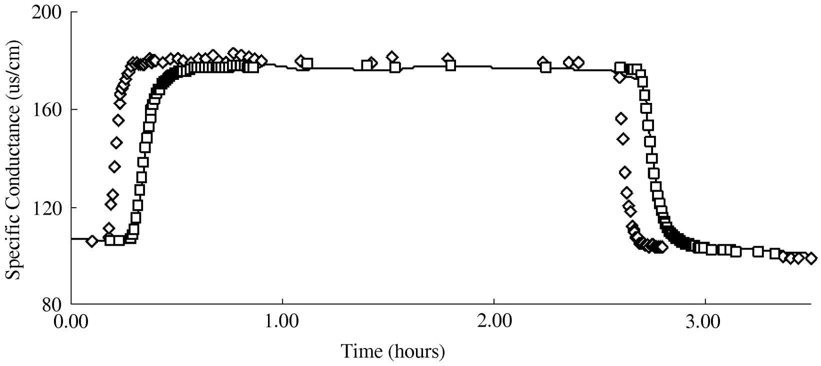

Figure 3. OTIS-P model simulation fit to LTC 4 August 2004 tracer curve

plot and one-way analysis of variance (ANOVA). Correlation coefficients were computed between measured discharge and the hyporheic flow variables. A total of 136 data sets (tracer curve runs) from 20 other field studies were obtained from the literature for comparison with our results (Table III). Ranges for basic variables and OTIS-P outputs from this study, work by others in coarse bed streams, and work by others in streams with beds finer than gravel were compared using scatter plots. In addition, the effects of natural and manmade flow obstructions (baffles, large wood formations, dense vegetation) on F_{med200} were considered by examining regression relationships between F_{med200} and the only component of F_{med200} that is independent of transient storage, i.e. the mean advective stream velocity u. The hypothesis that the relationship between F_{med200} and u would be different for obstructed and unobstructed streams was tested by comparing regression coefficients for linear regression of log F_{med200} versus log u. The significance of differences in these coefficients was determined using Monte Carlo analysis. Observations were randomly assigned to obstructed and unobstructed categories and linear regression was conducted on the resulting data. The difference between the slope of the line for obstructed sites and the slope of the line for unobstructed sites was computed. The procedure was repeated 5000 times, and the number of times the slope difference exceeded the slope difference based on real data was counted.

Hyporheic relative to total storage

The quantity q_s ($\mathrm{m^3\ m^{-1}\ s^{-1}}$), the average exchange of water through storage zones per unit length of stream (i.e. total storage exchange), was computed by

$$q_s = \alpha A \tag{5}$$

and the average flux of water through all storage zones Q_s ($\mathrm{m^3\ s^{-1}}$) was obtained by multiplying q_s by the reach length L:

$$Q_s = \alpha A L \tag{6}$$

Dividing Q_s by A_s gives the average rate at which stream water is exchanged for water in the storage zones q'_s ($\mathrm{m\ s^{-1}}$):

$$q'_s = \frac{Q_s}{A_s} = \frac{\alpha A L}{A_s} \tag{7}$$

Using the VHG ($\Delta h/\Delta d$), where $\Delta d = 0{\cdot}25$ m, and streambed conductivity K_{sp}, the local rate at which stream water is exchanged for water in the hyporheic zone $q_h(x)$ ($\mathrm{m\ s^{-1}}$) may be computed:

$$q_h(x) = -K_{\mathrm{sp}}\frac{\Delta h}{\Delta d} \tag{8}$$

where x is location along the stream reach. The reach-average specific discharge between the stream and the groundwater q'_h ($\mathrm{m\ s^{-1}}$) may be obtained by integration:

$$q'_h = \frac{\int_0^L |q_h(x)|\mathrm{d}x}{L} = \frac{\sum_{i=1}^{N} |q_h(x_i)|\ell(x_i)}{L} \tag{9}$$

Table III. Published field studies containing measurements of transient storage in streams. Entries represent the number of tracer tests reported

Streambed material type: Other	Streambed material type: Sand and finer	Reference
	1	Battin *et al.* (2003)
4		Bencala (1984)
3		Bencala and Walters (1983)
8		Bencala *et al.* (1990)
	4	Bohlke *et al.* (2004)
4		Broshears *et al.* (1993)
4		Broshears *et al.* (1996)
	1	Chapra and Wilcock (2000)
4		D'Angelo *et al.* (1993)
	8	Ensign and Doyle (2005)
37		Hall *et al.* (2002)
20		Hart *et al.* (1999)
10		Harvey and Fuller (1998)
2		Harvey *et al.* (1996)
5		Harvey *et al.* (2003)
	9	Jin and Ward (2005)
3		Lautz *et al.* (2006)
5	1	Morrice *et al.* (1997)
2		Mulholland *et al.* (1997)
	1	Mutz and Rohde (2003)
111	25	Total

where N is the number of piezometers installed into the stream, and $\ell(x_i)$ is the length of stream segment represented by piezometer i. The term 'specific' discharge is used because it is a measure of the rate at which water is exchanged at the streambed–surface water interface. Since we do not know the actual hyporheic flow paths, we assume that hyporheic flux is equal or strongly related to flow across the bed surface. Accordingly, the ratio q'_h/q'_s is an indicator of the importance of hyporheic storage relative to total storage.

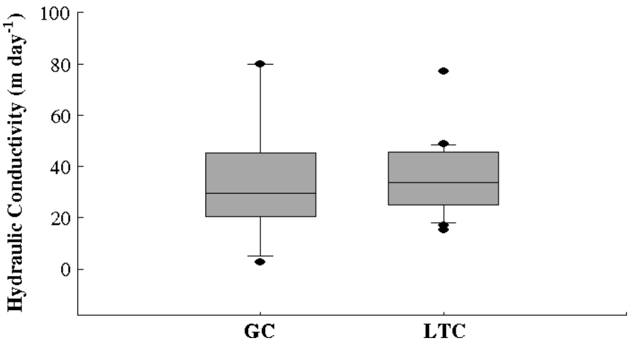

Figure 4. Streambed hydraulic conductivity obtained via falling-head tests in piezometers for GC and LTC. Central lines in boxes are medians, upper box boundaries are 75th percentiles, lower boundaries are 25th percentiles, and whiskers represent the 90th (upper) and 10th (lower) percentiles. Plotted points are outliers

RESULTS

Median bed material size at GC was eight times greater than at LTC. Sands from the two sites were about the same size, but GC sediments were bimodally distributed, with a component of gravel <30 mm (Figure 2). Hydraulic conductivity values obtained via falling-head tests were not significantly different between streams (one-way ANOVA, $F = 0{\cdot}012$; $p = 0{\cdot}9$, where p is the probability that this great a difference could occur due to chance alone): median conductivity for GC was 29·7 m day^{-1} (number of observations $n = 17$) and LTC was 33·6 m day^{-1} ($n = 22$) (Figure 4). Measured vertical water surface elevation differences inside and outside piezometers ranged from 2 to 22 mm, and the reach mean of the absolute values of VHGs ranged from 0·03 to 0·06, which compares with means ranging from 0·02 to 0·18 reported for the top 20 cm of a stream bed of coarser (~0·6 mm) sand (Battin *et al*., 2003). Piezometric observations typical of all experiments show both negative and positive vertical gradients within the study reaches, suggesting complex patterns of hyporheic inflow and outflow (Figure 5). Preliminary observations of piezometers placed along transects placed perpendicular to the stream indicated lateral hydraulic gradients were insignificant ($\leq \pm 0{\cdot}008$ m m^{-1}). Furthermore, based on measurements of in-channel discharge at the upstream and downstream ends of the study reaches, net lateral inflows and outflows during the dye studies were negligible.

Simulated time versus concentration curves generated using OTIS-P fit the observed conductivity data well based on linear regression of predicted versus actual values (slope: 0·99; intercept: 2·38 µs cm^{-1}; $r^2 = 0{\cdot}99$) and root-mean-square error, RMSE of 1·98 (Figure 3). Discharges ranged from 23 to 53 l s^{-1} and mean flow velocities from 0·02 to 0·17 m s^{-1} (Table IV). The lowest

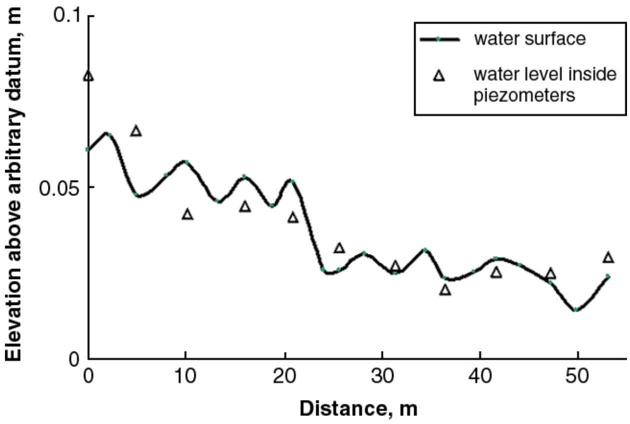

Figure 5. Water surface profile and water levels inside piezometers measured at Goodwin Creek on 6 June 2004

velocities occurred during experiments in which flow obstructions were present. Higher dispersion coefficients D were generally associated with higher mean flow velocities, and D was positively correlated with discharge Q ($r = 0{\cdot}69$). Values for storage exchange coefficient α, advective flow cross-sectional area A and storage zone area A_s are listed in Table IV. Values for α were not correlated with Q. The Damkohler numbers for the experiments ranged from 0·1 to 5·9 and averaged 1·9, indicating a valid experimental setup (Wagner and Harvey, 1997).

The two study streams responded differently to the flow obstructions (Table V). The ratio A_s/A and the main channel travel distance L_s decreased after beaver dam construction at GC, but increased after artificial log

Table IV. Outputs from OTIS-P model simulations

Stream	Date (2004)	Obstruction	Q (l s^{-1})	u (m s^{-1})	D (m^2 s^{-1})	α (s^{-1})	A (m^2)	A_s (m^2)	DaI
GC	26 May	No	39	0·09	0·43	0·0008	0·41	0·13	0·6
	7 June	No	53	0·12	0·94	0·0001	0·46	0·45	0·1
	14 July	No	53	0·08	0·47	0·0004	0·64	1·04	0·8
	30 July	Full beaver dam	23	0·02	0·04	0·0005	0·94	0·26	1·7
	13 August	Partial beaver dam	25	0·06	0·39	0·0007	0·41	0·16	1·0
LTC	22 July	No	51	0·17	3·50	0·0005	0·30	0·03	0·2
	4 August	Artificial underflow jam	36	0·13	2·55	0·0001	0·28	0·88	0·2

DOI: 10.1002/hyp

Table V. Metrics used to describe response of conservative tracer to storage processes

Stream	Obstruction	A_s/A	L_s (m)	R_h (s m^{-1})	F_{med200}	q'_s (m s^{-1})	q'_h (10^{-5} m s^{-1})	$100 \times q'_h/q'_s$ (%)
GC	No	0·31	116	3·30	0·09	0·141	3·15	0·02
	No	0·99	923	8·41	0·03	0·007	1·79	0·26
	No	1·63	225	19·67	0·15	0·014	—	—
Mean (GC)	No	1·0	422	10·0	0·09	0·054	2·32	0·10
	Full beaver dam	0·27	48	11·51	0·15	0·114	1·69	0·01
	Partial beaver dam	0·38	85	6·25	0·15	0·119	1·85	0·02
Mean (GC)	Yes	0·3	67	8·9	0·15	0·116	1·77	0·02
LTC	No	0·11	354	0·65	0·02	0·282	2·79	0·01
	Artificial underflow jam	3·19	1200	25·13	0·04	0·002	1·05	0·49

placement at LTC by factors of 3 (A_s/A) and 6·3 (L_s). The mean hydraulic retention factor R_h decreased about 10% after beaver dam construction at GC, but increased by a factor of 40 after placement of the flow obstruction at LTC. On the other hand, the fraction of median travel time due to storage F_{med} approximately doubled at both sites following flow obstruction and varied in a linear fashion with the Darcy friction factor f. It should be noted that variations in metric response may be partially due to parameter uncertainty relative to each experiment, as it is impossible to guarantee DaI = 1·0 prior to each experiment.

During the 30 July 2004 test at GC, obstruction by the beaver dam produced non-advective flow and invalidated the use of average bed slope to derive Darcy f. If this data point is omitted from the regression, then variation in Darcy f explains 94% of the variation in F_{med200} ($p < 0{\cdot}001$, Figure 6). Computed reach averaged specific discharge q'_h averaged $2{\cdot}05 \times 10^{-5}$ m s^{-1} for all tests, and varied little, with a range of $1{\cdot}7 \times 10^{-5}$ to $3{\cdot}2 \times 10^{-5}$ m s^{-1} for GC and $1{\cdot}1 \times 10^{-5}$ to $2{\cdot}8 \times 10^{-5}$ m s^{-1} for LTC. Assuming an uncertainty of 2 mm for pressure head observations, the uncertainty associated with the corresponding hyporheic flux potential q'_h ranges from approximately $2{\cdot}8 \times 10^{-6}$ m s^{-1} to $4{\cdot}1 \times 10^{-6}$ m s^{-1} at GC and from $2{\cdot}3 \times 10^{-6}$ m s^{-1} to $3{\cdot}9 \times 10^{-6}$ m s^{-1} at LTC. Additional, but likely small, uncertainty arises due to uncertainty associated with estimates of the hydraulic conductivity K_{sp}. The ratio q'_h/q'_s ranged from 0·01% to 0·49%, and was lower at GC following beaver dam construction, but increased at LTC following placement of flow obstruction (Table V).

Of the 136 data sets found in the literature, 111 were for streams with beds coarser than sand. We were not able to find any reports of simultaneous measurements of bed hydraulic conductivity and VHG for sandy streams in the literature. Values for stream discharge and bed slope observed in this study were intermediate to values reported by others for sand-bed streams (Figure 7). As might be expected, reported gradients for sandy streams tended to be up to two orders of magnitude smaller and mean stream velocity one order of magnitude smaller than for coarser bedded streams. Parameters describing total transient storage along the two streams

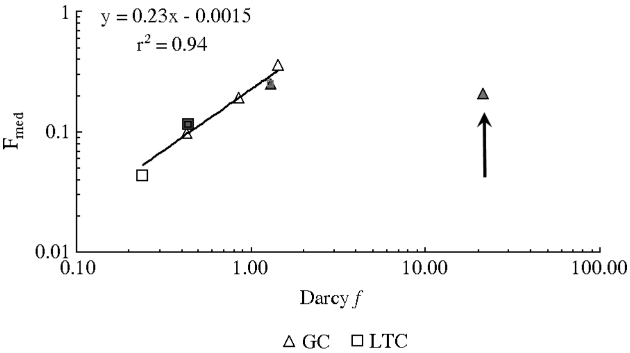

Figure 6. Fraction of median travel time due to storage F_{med200} versus the Darcy friction factor f for GC and LTC. Solid symbols represent experiments in which flow obstructions were present within the study reach. Point marked with arrow indicates case where downstream beaver dam created backwater, invalidating use of average bed slope to compute Darcy f

studied were within ranges reported by others, but values for the storage zone exchange coefficient α and mean stream velocity u were confined to the lower ends of reported ranges (Figure 7b). Values of A_s/A for our sites averaged $0{\cdot}98 \pm 1{\cdot}11$, which lies closer to the mean for all coarse-bed data ($0{\cdot}47 \pm 0{\cdot}64$) than for sand streams ($0{\cdot}36 \pm 0{\cdot}22$) (Figure 7c). Values of α and A_s/A were not correlated with discharge ($r^2 < 0{\cdot}01$).

F_{med200} and R_h are measures of total transient storage that are useful for intersite comparisons. Values of both metrics for our tests were within the range of values reported by others for sand-bed streams. Published values of F_{med200} and R_h for sand-bed streams were found in the upper part of the range reported for coarser bedded streams (Figure 7d), indicating that these streams tended to have higher levels of transient storage. When our data were pooled with the published data, F_{med200} and R_h were both inversely proportional to discharge ($p < 0{\cdot}02$). As expected, based on functional relationship (Equation (4)), F_{med200} was also inversely proportional to mean stream velocity u ($r = -0{\cdot}41$, $p < 0{\cdot}001$), with only 5 of 10 values >10% when velocity exceeds 0·1 m s^{-1} and none >27% (Figure 8). Stream reaches for which investigators reported natural or artificial obstructions (usually 'debris dams') tended to have lower values of F_{med200} for a given value of u (Figure 8), but regression coefficients were not significantly different ($p > 0{\cdot}05$).

DOI: 10.1002/hyp

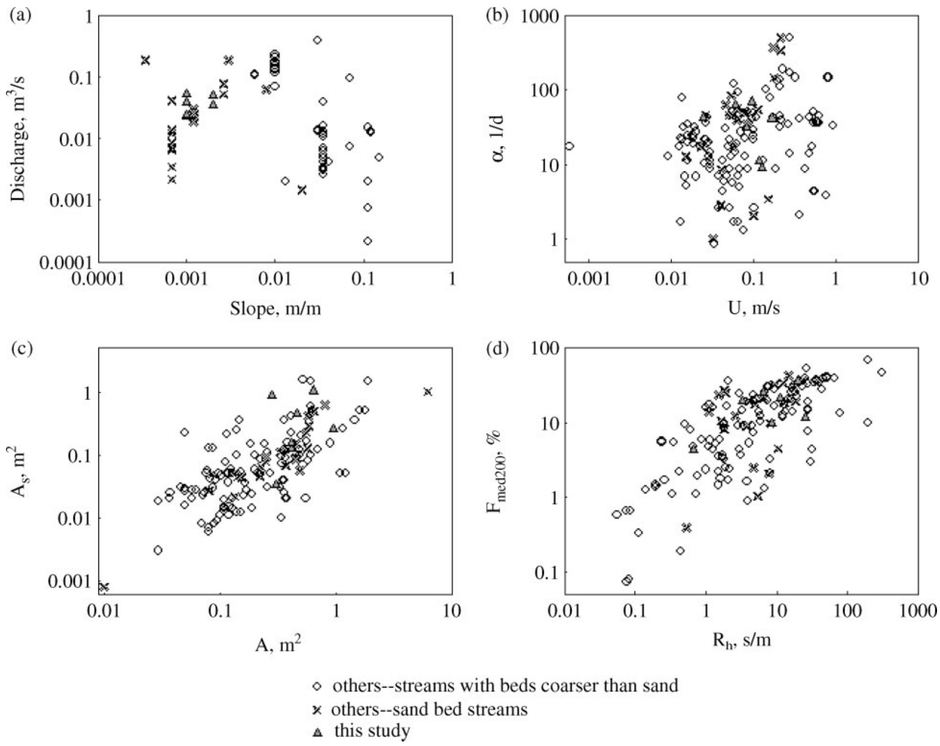

Figure 7. Comparison of results of this study with data compiled from literature (Table V): (a) discharge and stream bed slope, (b) storage exchange coefficient and mean stream velocity; (c) cross-sectional area of storage zones and stream cross-sectional area; (d) fraction of median travel time due to storage and hydraulic retention factor

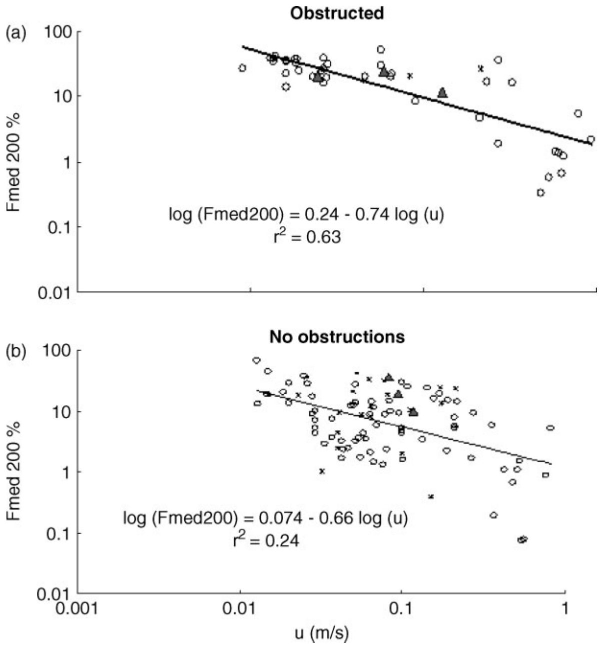

Figure 8. Fraction of median travel time due to transient storage (standardized to 200 m long reach) F_{med200} as a function of mean stream velocity u. Values of u obtained by dividing the measured discharge by flow cross-sectional area obtained from OTIS-P inverse modelling. (A) Streams obstructed by large wood, debris dams, manmade baffles or other solid objects. (B) Streams without obstruction. Regression lines are fitted to all data

DISCUSSION

Here, we present a method to quantify hyporheic and total storage exchange for a small stream reach. The measurement of VHG and streambed conductivity allows estimation of the average velocity of flow across the streambed interface, which can be compared with the total storage exchange rate. The proposed method does not quantify hyporheic zone flow paths as others have done (e.g. Kasahara and Wondzell, 2003; Gooseff *et al*., 2006), but it is consistent with concepts underlying the widely used one-dimensional transient storage model (Runkel, 1998). A shortcoming of this research was that piezometers were only installed to one depth in the streambed. Future research that attempts to quantify hyporheic exchange should install piezometers at different depths to explore vertical heterogeneity.

The proposed method was used to study transient storage in two small, sand-bed streams. Determination of hydraulic conductivity K_{sp} is an essential part of the method, and we measured values similar to those reported by Landon *et al*. (2001) for sandy streams. Despite the presence of gravel ($\sim$30 mm) at GC, the between-streams difference in K_{sp} distributions was small (Figure 3), which was likely due to the interstitial spaces in the bed matrix at GC being filled with medium sands (Hancock *et al*., 2001).

The OTIS-P output parameters for the GC tests displayed considerable temporal variability. This variability may be due to uncertainty in parameter estimates. OTIS-P generates values for several parameters that produce the best model fit to observed data based on minimizing the sums of squares, but it does not give parameter estimates from nearly equal model fits using a maximum likelihood approach. This is a shortcoming in this standard tool. To address this issue, we ensured that the model-predicted

DOI: 10.1002/hyp

cross-sectional flow area and dispersion coefficients were realistic for the experimental conditions. However, we realize that this approach did not fully eliminate the problem.

Another potential source of temporal variation in model outputs for the same reach may have been changes in subsurface flow paths caused by high flow events that occurred between experiments. The experiments performed without flow obstruction on 7 June 2004 and 14 July 2004 were conducted with similar discharges, but an upstream gauge recorded six flow events between these two tests, all with peaks >5000 l s^{-1}, or about 100 times the flow rate that was measured during the tracer studies. Even high-frequency flow events in the GC study reach generate adequate shear stress to entrain all bed materials, and bed disturbance that appears after high flows is manifest by sand waves and deposits of sand and gravel that are too soft to walk on. Although discharges were similar for the 7 June 2004 and 14 July 2004 tests, values of L_s, R_h, and F_{med200} all indicate greater hydraulic retention on the later date (Table V). Greater hydraulic retention on 14 July 2004 was associated with 25% greater flow width, 10% greater flow depth, 30% lower mean velocity, and a Darcy friction factor that was more than three times greater. Unfortunately, streambed hydraulic conductivity was not measured on 14 July 2004, when elevated retention occurred. At LTC, bed grain size distribution was more uniform and there was less variability in hydraulic conductivity and OTIS-P output parameters. Metrics A_s/A, R_h, and F_{med} all increased with the presence of flow obstruction, indicating a positive association between obstruction and total storage.

A compilation of tracer study results from mountain streams (Runkel, 2002) showed that the influence of storage processes tends to decrease as stream velocity increases, and our data and our compilation of data from the literature showed that F_{med200} and R_h decline with increasing velocity. At GC, the lowest mean velocity occurred when the fully intact beaver dam was present. Previous studies have also shown that debris dams or flow obstructions increased hyporheic zone flow in a stream in the intermountain western USA (Lautz *et al.*, 2006), but this pattern was not observed in these low-gradient, sand-bed streams. However, water pooling behind the flow obstructions may have created hyporheic flow paths with time-scales that were too long to be captured by the stream-tracer approach. Evidently, transient storage in small, sand-bed streams is dominated by surface storage zones, and the size and influence of these zones is heavily dependent on flow obstructions, such as large wood and debris dams (Ensign and Doyle, 2005). Despite considerable scatter in the data and attendant lack of statistically significant differences between the regression coefficients, the regression formula for obstructed streams in Figure 8 predicts values of F_{med200} that are 1·7 times as great for a velocity of 0·1 m s^{-1} as the line for unobstructed streams.

CONCLUSIONS

The proposed methodology for separating hyporheic and surface storage may prove to be a useful tool when applied to other stream systems, especially for understanding existing storage mechanisms. OTIS-P output parameters fell within the ranges reported for streams with beds much coarser than sand, despite the fact that coarser beds are assumed to provide higher levels of hyporheic storage and exchange. The primary mechanism of storage occurs through surface as opposed to hyporheic pathways, primarily due to the streams' low gradients and fairly uniform bed composition. Hyporheic exchange accounted for a small percentage (i.e. $<0{\cdot}5\%$) of total storage exchange in the sand-bed streams examined in this study, and this fraction did not increase due to flow obstructions. Furthermore, the findings suggest that stream–hyporheic zone exchange in lowland streams may not be significant in terms of biogeochemical processing. This conclusion may impact interest in stream–hyporheic zone exchange as part of an integrated approach to catchment management. On the other hand, the fraction of median travel time due to storage, F_{med}, approximately doubled at both sites following flow obstruction and varied in a linear fashion with the Darcy friction factor f. A compilation of published data showed that observed $F_{med200} < 30\%$ for stream reaches with mean velocity $>0{\cdot}1$ m s^{-1} and $F_{med200} < 10\%$ for reaches with mean velocity $>0{\cdot}2$ m s^{-1}. Removal of roughness elements, such as large wood and debris dams, is detrimental to processes dependent upon transient surface storage, but not hyporheic storage, in small, sand-bed streams.

ACKNOWLEDGEMENTS

Reviews of an earlier version of this paper by Michael Gooseff, Martin Doyle, and Judson Harvey are gratefully acknowledged. Assistance with fieldwork was provided by John Massey, Calvin Vick, Landon Lee, Robbie Kroger, and Tim Sullivan. Michael Ursic assisted with manuscript preparation.

REFERENCES

Adams FA. 2000. *Geologic investigation report: Yalobusha River subwatershed, Little Topashaw Creek stream corridor rehabilitation, Chickasaw County, Mississippi*. US Department of Agriculture, Natural Resources Conservation Service: Jackson, MS (unpublished).

Battin TJ, Kaplan LA, Newbold JD, Hendricks SP. 2003. A mixing model analysis of stream solute dynamics and the contribution of a hyporheic zone to ecosystem function. *Freshwater Biology*, **48**: 1–20.

Baxter C, Hauer FR, Woessner WW. 2003. Measuring groundwater-stream water exchange: new techniques for installing minipiezometers and estimating hydraulic conductivity. *Transactions of the American Fisheries Society* **132**: 493–502.

Bencala KE. 1984. Interactions of solutes and streambed sediment 2. Dynamic analysis of coupled hydrologic and channel processes that determine solute transport. *Water Resources Research* **20**: 1804–1814.

Bencala KE. 2005. Hyporheic exchange flows. In *Encyclopedia of Hydrological Sciences*, Vol. 3, Part 10, Anderson MG, McDonnell JJ (eds). Wiley: Chichester; 1733–1740.

Bencala KE, Walters RA. 1983. Simulation of solute transport in a mountain pool-and-riffle stream: a transient storage model. *Water Resources Research* **19**: 718–724.

Bencala KE, McKnight DM, Zellweger GW. 1990. Characterization of transport in an acidic and metal-rich mountain stream based on a lithium tracer injection and simulations of transient storage. *Water Resources Research* **26**: 989–1000.

Bohlke JK, Harvey JW, Voytek MA. 2004. Reach-scale isotope tracer experiment to quantify denitrification and related processes in a nitrate-rich stream, mid-continent United States. *Limnology and Oceanography* **49**: 821–838.

Boulton AJ. 2000. The functional role of the hyporheos. *Verh. Internat. Verein. Limnol.*, **27**: 51–63.

Boulton AJ, Depauw S, Marmonier P. 2002. Hyperheic dynamics in degraded rural stream carrying a 'sand slug'. *Verhandlungen der Internationalen Vereinigung für Theoretische und Angewandte Limnologie* **28**: 120–124.

Broshears RE, Bencala KE, Kimball BA, McKnight DM. 1993. *Tracer-dilution experiments and solute-transport simulations for a mountain stream, Saint Kevin Gulch, Colorado*. USGS Water-Resources Investigations Report 92–4081, USGS, Denver; 1–18.

Broshears RE, Runkel RL, Kimball BA, McKnight DM, Bencala KE. 1996. Reactive solute transport in an acidic stream: experimental pH increase and simulation of controls on pH, aluminum, and iron. *Environmental Science and Technology* **30**: 3016–3024.

Chapra SC, Wilcock RJ. 2000. Transient storage and gas transfer in lowland stream. *Journal of Environmental Engineering* **126**: 708–712.

Cleven E-J, Meyer EI. 2003. A sandy hyporheic zone limited vertically by a solid boundary. *Archiv für Hydrobiologie* **157**: 267–288.

D'Angelo DJ, Webster JR, Gregory SV, Meyer JL. 1993. Transient storage in Appalachian and Cascade mountain streams as related to hydraulic characteristics. *Journal of the North American Benthological Society* **12**: 223–235.

Ensign SH, Doyle MW. 2005. In-channel transient storage and associated nutrient retention: evidence from experimental manipulations. *Limnology and Oceanography* **50**: 1740–1751.

Fox GA. 2004. Evaluation of a stream–aquifer analysis test using analytical solutions and field data. *Journal of the American Water Resources Association* **40**: 755–763.

Gooseff MN, LaNier J, Haggerty R, Kokkeler K. 2005. Determining in-channel (dead zone) transient storage by comparing solute transport in a bedrock channel-alluvial channel sequence, Oregon. *Water Resources Research* **41**: W06014. DOI: 10·1029/2004WR003513.

Gooseff MN, Anderson JK, Wondzell SM, LaNier J, Haggerty R. 2006. A modelling study of hyporheic exchange pattern and the sequence, size, and spacing of stream bedforms in mountain stream networks, Oregon, USA. *Hydrological Processes* **20**: 2443–2457.

Hall Jr RO, Bernhardt ES, Likens GE. 2002. Relating nutrient uptake with transient storage in forested mountain streams. *Limnology and Oceanography* **47**: 255–265.

Hancock P, Boulton A, Raine A. 2001. Surface–subsurface hydrological connectivity in sand-bed streams. In *Proceedings of the Third Australian Stream Management Conference*, Rutherford I, Sheldon F, Brierley G, Kenyon C (eds), 27–29 August 2001, Brisbane, Australia; 259–264.

Hart DR, Mullholland PJ, Marzolf ER, DeAngelis DL, Hendricks SP. 1999. Relationships between hydraulic parameters in a small stream under varying flow and seasonal conditions. *Hydrological Processes* **13**: 1497–1510.

Harvey JW, Fuller CC. 1998. Effect of enhanced manganese oxidation in the hyporheic zone on basin-scale geochemical mass balance. *Water Resourses Research* **34**: 623–636.

Harvey JW, Wagner BJ. 2000. Quantifying hydrologic interactions between streams and their subsurface hyporheic zone. In *Streams and Ground Waters*, Jones JA, Mulholland PJ (eds). Academic Press: San Diego, CA; 1–43.

Harvey JW, Wagner BJ, Bencala KE. 1996. Evaluating the reliability of the stream tracer approach to characterize stream–subsurface water exchange. *Water Resources Research* **32**: 2441–2451.

Harvey JW, Conklin MH, Koelsch RS. 2003. Predicting changes in hydrologic retention in an evolving semi-arid alluvial stream. *Advances in Water Resources* **26**: 939–950.

Hauer FR, Dahm CN, Lamberti GA, Stanford JA. 2003. Landscapes and ecological variability of rivers in North America: factors affecting restoration strategies. In *Strategies for Restoring River Ecosystem: Sources of Variability and Uncertainty in Natural and Managed Systems*, Wissmar RC, Bisson PA (eds). American Fisheries Society: Bethesda, MD; 81–105.

Hendricks SP. 1993. Microbial ecology of the hyporheic zone: a perspective integrating hydrology and biology. *Journal of the North American Benthological Society* **12**: 70–78.

Jin HS, Ward GM. 2005. Hydraulic characteristics of a small Coastal Plain stream of the southeastern United States: effects of hydrology and season. *Hydrological Processes* **19**: 4147–4160.

Kasahara T, Hill AR. 2006. Hyporheic exchange flows induced by constructed riffles and steps in lowland streams in southern Ontario, Canada. *Hydrological Processes* **20**: 4287–4305.

Kasahara T, Wondzell SM. 2003. Geomorphic controls on hyporheic exchange flow in mountain streams. *Water Resources Research* **39**: 1005. DOI: 10·1029/2002WR001386.

Landon MK, Rus DL, Harvey FE. 2001. Comparison of instream methods for measuring hydraulic conductivity in sandy streambeds. *Ground Water* **39**: 870–885.

Lautz LK, Siegel DI, Bauer RL. 2006. Impact of debris dams on hyporheic interaction along a small semi-arid stream. *Hydrological Processes* **20**: 183–196.

Morrice JA, Valett HM, Dahm CN, Campana ME. 1997. Alluvial characteristics, groundwater–surface exchange and hydrological retention in headwater streams. *Hydrological Processes* **11**: 253–267.

Mulholland PJ, Marzolf ER, Webster JR, Hart DR, Hendricks SP. 1997. Evidence that hyporheic zones increase heterotrophic metabolism and phosphorus uptake in forest streams. *Limnology and Oceanography* **42**: 443–451.

Munn NL, Meyer JL. 1990. Habitat-specific solute retention in two small streams: an intersite comparison. *Ecology* **71**: 2069–2082.

Mutz M, Rohde A. 2003. Processes of surface–subsurface water exchange in a low energy sand-bed stream. *International Review of Hydrobiology* **88**: 290–303.

Palmer MA. 1993. Experimentation in the hyporheic zone: challenges and prospectus. *Journal of the North American Benthological Society* **12**: 84–93.

Runkel RL. 1998. *One-dimensional transport with inflow and storage (OTIS): a solute transport model for streams and rivers*. US Geological Survey Water-Resources Investigations Report 98–4018.

Runkel RL. 2002. A new metric for determining the importance of transient storage. *Journal of the North American Benthological Society* **21**: 529–543.

Runkel RL, Bencala KE. 1995. Transport of reacting solutes in rivers and streams. In *Environmental Hydrology*, Singh VP (ed.). Kluwer: Dordrecht; 137–163.

Runkel RL, Broshears RE. 1991. *One-dimensional transport with inflow and storage (OTIS)—a solute transport model for small stream: Boulder, Colo*. University of Colorado, CADSWES Technical Report 91–10.

Shields Jr FD, Gippel CJ. 1995. Prediction of effects of woody debris removal on flow resistance. *Journal of Hydraulic Engineering* **121**: 341–354.

Simon A. 1989. A model of channel response in disturbed alluvial channels. *Earth Surface Processes and Landforms* **14**: 11–26.

Stofleth JM, Shields Jr FD, Fox GA. 2004. Organic carbon concentrations in hyporheic zone sediments: a tool for measuring stream integrity. In *Proceedings of the 2004 World Water and Environmental Resources Congress: Critical transitions in Water and Environmental Resources*, Sehlke G, Hayes DF, Stevens DK (eds). American Society of Civil Engineers: Reston, VA (CD-ROM).

Triska FJ, Kennedy VC, Avanzino RJ, Zellweger GW, Bencala KE. 1989. Retention and transport of nutrients in a third-order stream in northwestern California: hyporheic processes. *Ecology* **70**: 1893–1905.

Triska FJ, Duff JH, Avanzino RJ. 1993. Patterns of hydrological exchange and nutrient transformations in the hyporheic zone of a gravel-bottom stream: examining terrestrial-aquatic linkages. *Freshwater Biology* **29**: 259–274.

Wagner BJ, Harvey JW. 1997. Experimental design for estimating parameters of rate-limited mass transfer: analysis of stream tracer studies. *Water Resources Research* **33**: 1731–1741.

Wallerstein N, Thorne CR, Doyle MW. 1997. Spatial distribution and impact of large woody debris in northern Mississippi. In *Management of Landscapes Disturbed by Channel Incision, Stabilization, Rehabilitation, and Restoration*, Wang SSY, Langendoen EJ, Shields Jr FD (eds). University of Mississippi: Mississippi; 145–150.

DOI: 10.1002/hyp